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CHANGES IN MORPHOLOGY AT THE MARGIN OF THE GREENLAND ICE SHEET (LEVERETT GLACIER), IN THE PERIOD 1943–1992: A QUANTITATIVE ANALYSIS

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ABSTRACT

Changes in ice-marginal morphology near Leverett glacier, a small outlet glacier at the western margin of the Greenland ice sheet, are determined from a photogrammetrical analysis. To be able to compare two datasets from subsequent years with measurements at different coordinates, kriging was used for interpolation. In this study the kriging standard error is used to evaluate the relative accuracy of the resulting maps. Aerial photographs of 1943, 1968 and 1985 were compared. In the period 1943–1968 an area of $0{\cdot}2 \times 10^6\,m^2$ was deglaciated. Approximately $1{\cdot}1 \times 10\,m^3$ of material is deposited in this area. The southern part of the deglaciated area is characterized by ice-cored moraines, while moraines without ice core were formed in the north. Differences in depositional products reflect differences in meltwater activity and probably ice-marginal thermal regime. During deglaciation a small proglacial sandur decreased in altitude by $3{\cdot}2 \pm 0{\cdot}1$ m. From the early 1970s Leverett glacier advanced over a previously deglaciated area. During this advance, small ice-marginal accumulations were incorporated and eroded by the advancing glacier. Erosion products were for a substantial part stored in the proglacial sandur. About $1{\cdot}2 \times 10^5\,m^2$ of the northern part of an ice-cored moraine complex decreased in altitude by $-3{\cdot}6 \pm 0{\cdot}1$ m from 1943 to 1968 and over $2{\cdot}7 \times 10^4\,m^2$ by $-2{\cdot}7 \pm 0{\cdot}1$ m during 1968–1985. The spatial patterns of altitude change were analysed in relation to topomorphological parameters as exposition and slope angle and areas occupied by lakes. The estimated energy used to melt the subsurface ice of the ice-cored moraine is $1{\cdot}4$–$2{\cdot}2\,W\,m^{-2}$ (1943–1968) and $1{\cdot}0$–$1{\cdot}6\,W\,m^{-2}$ (1968–1985). These values are 30–50 times larger than the geothermal heat flux. For the expected average debris concentration of the ice core (<10 per cent by volume) the deviation of the surface energy balance forced by climate change will be small and encompass an insignificant part of the total estimated energy used for melting.

KEY WORDS ice-marginal morphology; Greenland; geostatistics; moraines

INTRODUCTION

Historical fluctuations of the Greenland ice sheet are the object of a research project on the dynamics of the western ice-sheet margin. Research is focused on Leverett and Russell glacier in central west Greenland. Photogrammetrical analysis of Leverett glacier revealed fluctuations in glacier altitude and position of the ice margin (Van Tatenhove *et al.*, 1995). This paper concentrates on the changes in ice-marginal morphology during the period 1943–1992.

The aims of this paper are to quantify changes in ice-marginal morphology and to assess the problems in accurate description and analysis. Especially, the amount and rate of erosion and deposition within glacial and fluvioglacial systems during a period of glacier retreat (1943–1968) and a period of advance (1968–1992) are evaluated. Ice-marginal morphology can influence outlet glacier geometry. Local changes in ice thickness during the advance period can be explained by the overriding of frontal moraines (Van Tatenhove *et al.*, 1995). Knowledge of the sedimentation rates and the variability of ice-marginal sedimentation is essential for the interpretation of ice-marginal sedimentary dynamics in the past. This study gives information on the rate and scale of deposition and erosion in relation to short time (decades) fluctuations of the ice-sheet margin.

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The recent advance of large areas of the western margin of the Greenland ice sheet (Weidick *et al.*, 1992; Van Tatenhove *et al.*, 1995) for the first time enables an assessment of the change in ice-marginal geomorphology near an advancing ice sheet. This study is the first quantitative evaluation of these recent events.

SITE CHARACTERISTICS

The Leverett glacier in west Greenland is a small outlet glacier draining part of the main ice sheet. Its area encompasses approximately 20 km^2. In the atlas of glaciers in west Greenland, Leverett glacier is coded 1DG03002 (Weidick *et al.*, 1992). The part of the glacier surrounded by mountains is 7 km^2. These mountains have a maximum altitude of 582 m.a.s.l. (Figure 1). The area has a mean annual air temperature slightly lower than that of the nearby airport of Kangerlussuaq, $-5{\cdot}6 \pm 0{\cdot}2$°C. Mean annual precipitation is low (155 ± 7 mm) with a large variation in time and space. From mid-October till the end of April or beginning of May, a continuous snow cover is present with a thickness of 10–30 cm. The ground temperature in the proglacial area is below 0°C throughout the year, i.e. permafrost is continuous (Van Tatenhove and Olesen, 1994). Active layer thickness ranges between 0·1 and 2·5 m, depending on surface characteristics and topoclimatic position. The Greenland ice sheet reached its minimum position during the Hypsithermal. The location of this position is unknown, but could be up to 10–60 km east of the present margin (Weidick *et al.*, 1992; Letréguilly *et al.*, 1991). The main moraine phase before the Hypsithermal is known as the Ørkendalen phase (Ten Brink, 1975). Remnants of this phase are found 2 km west of the present ice margin (Figure 1) and consist of moraines and series of ice-marginal fluvioglacial terraces. An isolated morainic ridge in the centre of the Leverett basin is regarded as a product of the deglaciation following the Ørkendalen phase. From geoelectrical measurements the thickness of sediment in the centre of this basin is about 40–100 m (Van Tatenhove, 1993).

The most striking geomorphological feature is the central moraine complex in front of Leverett glacier which is the result of the advance after the Hypsithermal (Figure 1). This advance ended at the end of the last century. The modern age of the complex could be established by ^{14}C dating of organic material within fluvioglacial sediments in the southern part of the moraine complex (UtC-2536). Between this complex and the glacier margin is a small sandur.

The moraine complex consists of parallel ridges and furrows. The surface sediment on the ridges is generally a gravelly lag deposit due to the removal of fines by wind erosion. Surface sediments contain 19 per cent gravel, 61 per cent sand and 20 per cent silt and clay (five samples). Topographic low between ridges have a similar granulometry. On the west-facing slopes sandy aeolian deposits are common. Several closed depressions have been formed, especially in the northern part, and are occupied by lakes. The southern part of the complex is characterized by relatively deeply incised, fluvioglacial channels with bar deposits on the channel floors. In the south, ridges are composed of very fine silt (≈95 per cent silt, 'silt-cored moraines') under a veneer of gravel. Large parts of the complex are ice-cored. Close to the present ice margin, exposures give information on the characteristics of these ice-cored moraines. On top of an ice body of at least 4 m thickness, a sandy gravelly sediment cover of 2 m thickness is found. Within the ice body, small bands with a high debris concentration are folded, indicating a deformational history in the basal part of Leverett glacier. The base of the ice body is not exposed but will probably be on top of glacial or fluvioglacial sediments. At some sites organic material has been found in the sediment cover (see Figure 1). A date, UtC-2537 (2010 ± 80 a BP) indicates that this material has been incorporated by the advancing Leverett glacier after the Hypsithermal. The sandur between the moraine complex and Leverett glacier has fine gravelly to sandy surface deposits and some remnants of moraines in the centre. Remnants of multi-annual frost mounds are found on the east flank of the moraine complex. Within the fluvioglacial channels in the south, circular islands of gravels (diameter about 1 m) are common features. This type of patterned ground resembles stone pits (Washburn, 1979) and are here informally named 'stone boils' (after mudboils (Shilts, 1978)).

FIELD PROGRAMME AND PHOTOGRAMMETRICAL ANALYSIS

From large parts of the western margin of the Greenland ice sheet, aerial photographs are available from the

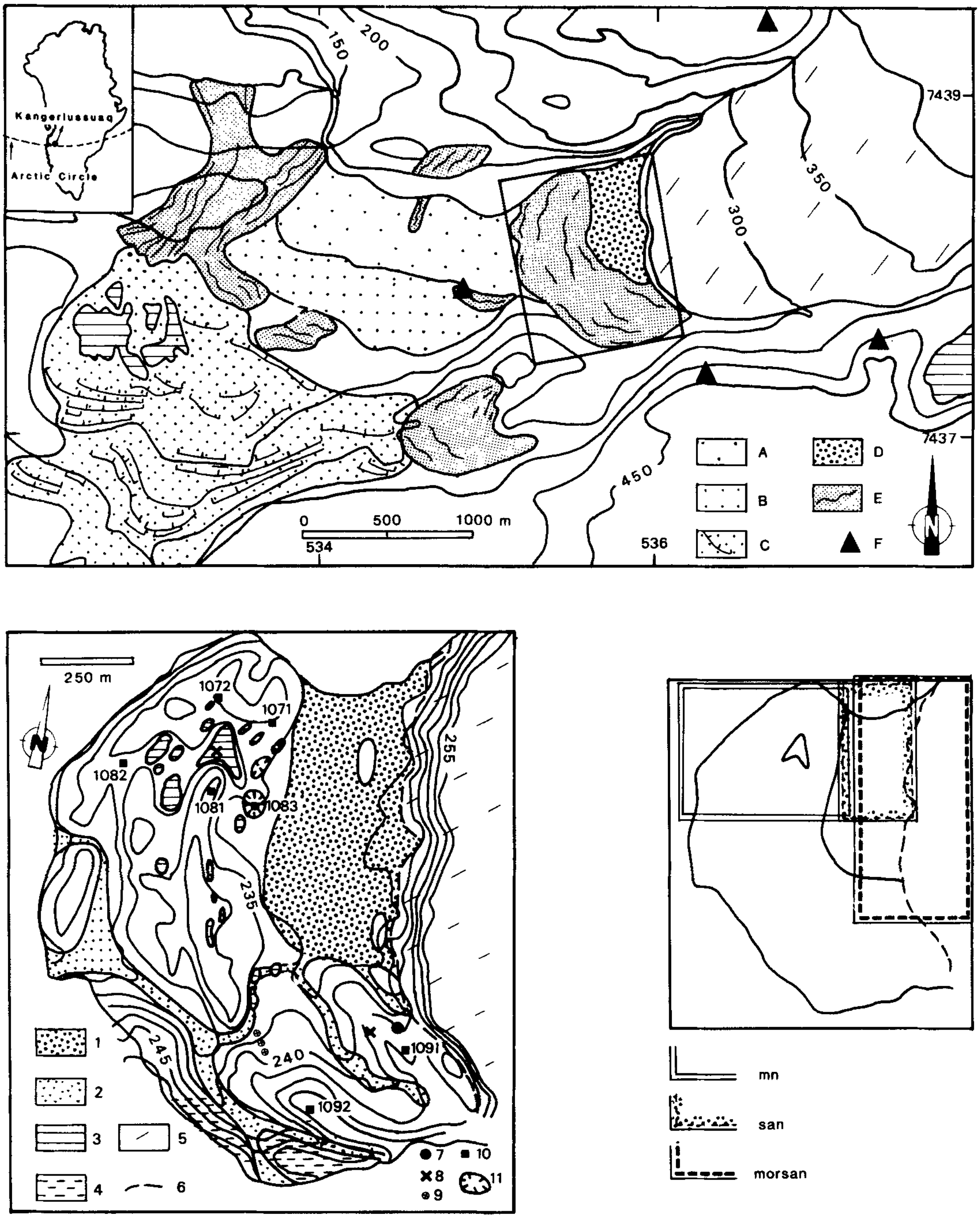

Figure 1. Location map of Leverett glacier. Legend of upper panel: A, bare sandur plain; B, vegetated sandur plain; C, ice-marginal terraces from the Ørkendalen period; D, sandur between glacier and moraine complex; E, moraine; F, geodetic control point. The lower panels show a map of the moraine complex (left) and an index of areas discussed in this paper (right). Legend: 1, sandur between glacier and moraine complex; 2, fluvioglacial deposits and inactive channel; 3, lake; 4, 'silt-cored' moraine; 5, glacier; 6, ice margin (1985); 7, location ground temperature measurement; 8, location water temperature measurement; 9, 'stone boil'; 10, point measured on moraines in 1992; 11, multi-annual frost mound. The names listed bottom right refer to Figures 4, 5 and 6 (moraine complex north 'mn'), Figure 9 (sandur, 'san') and Figures 8 and 10 (moraines on sandur 'morsan')

1940s, 1960s and 1980s. In the Leverett area, the aerial photographs were taken in 1943, 1968 and 1985 (for detailed information on the aerial photographs see Van Tatenhove *et al.* (1995)). After a short reconnaissance visit in 1991, a glacial-geological expedition visited Leverett glacier in 1992. During a six week period an inventory was made of ice-marginal morphological and sedimentological phenomena. Shallow soil temperatures were registered on three sites in the ice-margin area (Van Tatenhove and Olesen, 1994). Geodetical measurements were made with a distancemeter (AGA 220 Geodimeter) and theodilite (Wild T2). The position of the control points was based on the identification of points which are visible on each set of aerial photographs. The maximum standard deviations of the position of the control points (based on repeated levelling) are 3 cm in the horizontal directions, and 10 cm in elevation.

Fifteen points on ice-marginal moraines were measured (Figure 1). The positions of these points are less accurate than those of the control points. The maximum standard deviation in the position of points measured on moraines is 5 cm in the horizontal and 11 cm in the vertical.

Photogrammetrical measurements were carried out in an analytical stereoplotter (Zeiss Planicom C100) at the Department of Geodesy of the Technical University in Delft. In the photographs, the altitude of terrain elements could be measured with an accuracy of $\overline{dz} = 0{\cdot}86$ m, where $\overline{dz}$ is the square root of the sum of the squared residuals divided by the number of control points. The residuals refer to the differences between the terrain measurements and the photogrammetrical stereomodel (Van Tatenhove *et al.*, 1995). Altitude measurements on the moraines within the photographs were predominantly taken on important locations (ridges, slope breaks, valley between ridges, lake edges, moraine edges, channels, ice margin, etc.). Measurements were not performed on the same coordinates for each stereomodel. The data obtained from the photogrammetrical work consist of several hundreds of altitude values for each stereomodel.

From 18 June 1992 to 15 July 1992 soil temperature was measured in the sediment cover of an ice-cored moraine (for location see Figure 1). This cover is characterized by sandy gravelly material and during the period of measurement was moist to dry. The daily temperature amplitude was >0·15°C in the upper 0·85 m of the soil column. The estimated active layer thickness was 1·25–2·50 m, while the thermal diffusivity and the unfrozen thermal conductivity are estimated at $1{\cdot}32\ (\pm 0{\cdot}14) \times 10^{-6}\ m^2\ s^{-1}$ and $1{\cdot}8$–$4{\cdot}0\ W\ m^{-1}\ K^{-1}$, respectively (Van Tatenhove and Olesen, 1994). The estimated active layer thickness is confirmed by several observations in the moraine complex. Hence, sediment cover on the ice core of the moraine complex has similar values. Important in determining the amount of energy available for melting of the ice core is the ice/water content of the sediment layer on top of the ice core. In the summer of 1992 the upper 1·0 m was dry. Phase change at 1·0 m depth (temperature changed from −0·10 to +0·10°C) took place between 28 and 30 June (days 180–182). The time with nearly constant temperature close to the freezing point is short, indicating low moisture content at this depth (Figure 2).

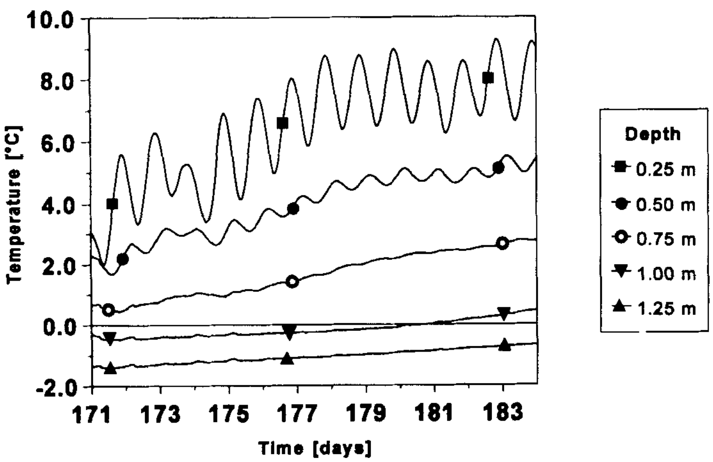

Figure 2. Ground temperature in the sediment cover of an ice-cored moraine in June–July 1992. Day 171 is 19 June

In the central part of the largest lake (Figure 1) within the moraine complex, temperatures were measured at the bottom of the lake at 4·5 m below lake level. The average water temperature in the period 8 to 11 July was 9·0 ± 0·3°C.

GEOSTATISTICAL DATA ANALYSIS

To be able to compare two datasets with altitude measurements at different coordinates, interpolation to a grid is necessary. It is important to evaluate the errors introduced by field measurements, photogrammetrical analysis and interpolation procedure. Using kriging as an interpolation technique provides a measure of the interpolated error at every interpolated point. To obtain an average altitude for a certain area, block kriging was performed. Because kriging takes account of the spatial structure of the data, an interpolation scheme can be constructed which suits the data best. In this study the kriging standard error is used to evaluate the relative accuracy of the resulting maps (altitude change and related maps). It will be shown that this relative accuracy is better than the accuracy of individual years due to a high correlation between interpolation errors of subsequent years.

The spatial structure of the measured moraine morphology and its changes in time can be described by means of variograms (Journel and Huijbrechts, 1978; Isaaks and Srivastava, 1989). In the usual intrinsic situation, the variogram of the spatial attribute Z (altitude) only depends on the distance and direction between two coordinates x and $x + h$:

$$\gamma(h) = \frac{1}{2}E[(Z(x) - Z(x+h))^2] \tag{1}$$

where E is the expected value.

Variograms are calculated with GSTAT (Pebesma, 1993). The 'nugget' is the intercept of the variogram with the y-axes, and is caused by measurement errors and short distance spatial variability. Consequently, the nugget should be equal to or greater than the variance of the residuals of the terrain measurements and the photogrammetrical stereomodel. The 'range' is the distance where the variogram remains essentially constant and is a measure of the distance of spatial correlation. The value of the variogram beyond the range is termed the 'sill'.

For describing the cross-relation between altitude measurements of two subsequent years i and j ($i, j = 1, 2$), normal cross-variograms are defined as:

$$\gamma_{ij}(h) = \frac{1}{2}E[(Z_i(x) - Z_i(x+h))(Z_j(x) - Z_j(x+h))] \tag{2}$$

Normal cross variograms can be expressed in terms of pseudo-cross-variograms, γ^{P}_{ij} according to Papritz *et al.* (1993):

$$\gamma_{ij}(h) = \frac{1}{2}\gamma^{\mathrm{P}}_{ij}(h) + \frac{1}{2}\gamma^{\mathrm{P}}_{ji}(h) - \gamma^{\mathrm{P}}_{ij}(0) \tag{3}$$

where the γ^{P}_{ij} is defined as (Papritz *et al.*, 1993):

$$\gamma^{\mathrm{P}}_{ij}(h) = \frac{1}{2}E[(Z_i(x+h) - Z_j(x))^2] \tag{4}$$

Pseudo-cross-variograms are used here as an intermediate step because measurements of altitude are on different coordinates for each year. This means that direct identification of the cross-variogram is not possible as can easily be seen from Equation 2. Independently modelled variograms are adjusted to obtain a linear model of coregionalization (Isaaks and Srivastava, 1989, p. 390). After evaluating the overall fit of auto- and cross-variogram models, it was verified that the resulting variograms satisfy the Cauchy–Schwartz inequalities given by (Goovaerts, 1994):

$$|\gamma_{ij}(h)| \leqslant \sqrt{\gamma_{ii}(h)\gamma_{jj}(h)} \text{ for all } h \tag{5}$$

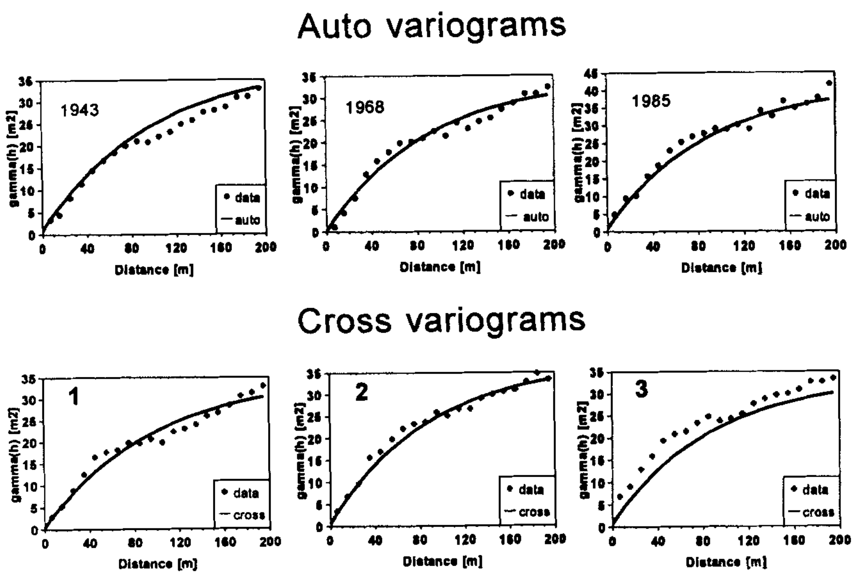

Figure 3. Upper: auto-variograms with fitted lines for the northern part of the moraine complex ('mn'). Variograms shown are 'mn 43 × 68' (1943), 'mn 68 × 85' (1968) and 'mn 68 × 85' (1985). Parameters of the variograms are given in Table I. Lower: cross-variograms and fitted variogram. 1, Cross-variogram for period 1943–1968 ('mn 43 × 68' in Table I); 2, period 1968–1985 ('mn 68 × 85'); 3, period 1943–1985 ('mn 43 × 85')

Table I. Parameters for the variograms and cross-variograms: exp = exponential model; sph = spherical model. Northern part of the moraine complex ('mn'), ice margin area in 1968 ('morsan') and proglacial sandur ('san')

	Best fit				Adjusted for cokriging				
	Model	Nugget	Sill	Range	Model	Nugget	Sill	Range	
Auto-variograms (number of datapoints)									
mn 43 (591)	exp	0·9	37·8	118	exp	0·90	37·8	100	mn 43 × 68
					exp	0·75	33·5	90	mn 43 × 85
mn 68 (402)	exp	0·2	34·2	93	exp	0·20	34·2	100	mn 43 × 68
					exp	0·20	34·2	90	mn 68 × 85
mn 85 (242)	exp	1·0	40·8	84	exp	1·00	40·8	90	mn 43 × 85
					exp	1·00	40·8	90	mn 68 × 85
san 43 (381)	sph	1·0	16·1	180	sph	1·00	16·1	180	san 43 × 68
san 68 (290)	sph	0·8	2·5	180	sph	0·80	2·5	180	san 43 × 68
					sph	0·80	2·5	180	san 68 × 85
san 85 (240)	sph	2·0	11·0	180	sph	1·40	3·2	180	san 68 × 85
morsan 68 (714)	exp	2·0	28·0	225					
Cross-variograms									
mn 43 × 68	exp	0·9	22·0	38	exp	0·4	35·0	100	
mn 68 × 85	exp	0·2	35·6	77	exp	0·6	37·0	90	
mn 43 × 85	exp	2·0	33·0	75	exp	0·8	33·0	90	
san 43 × 68	sph	13·0	6·3	180	sph	0·8	6·3	180	
san 68 × 85	sph	4·4	3·2	180	sph	1·4	3·2	180	

The study area in Figure 1 was divided into three areas ('mn', 'san' and 'morsan'). Examples of variograms are given in Figure 3. Models used (exponential and spherical) and parameter values are tabulated in Table I. The variogram models for individual years and the adjusted normal cross-variogram are used for ordinary block cokriging with a block size of 10 m ('mn' and 'morsan') and 25 m ('san'). Calculations are performed with GSTAT (Pebesma, 1993). Final results are the estimated altitude $Z^*(x)$ at grid x, the kriging variance

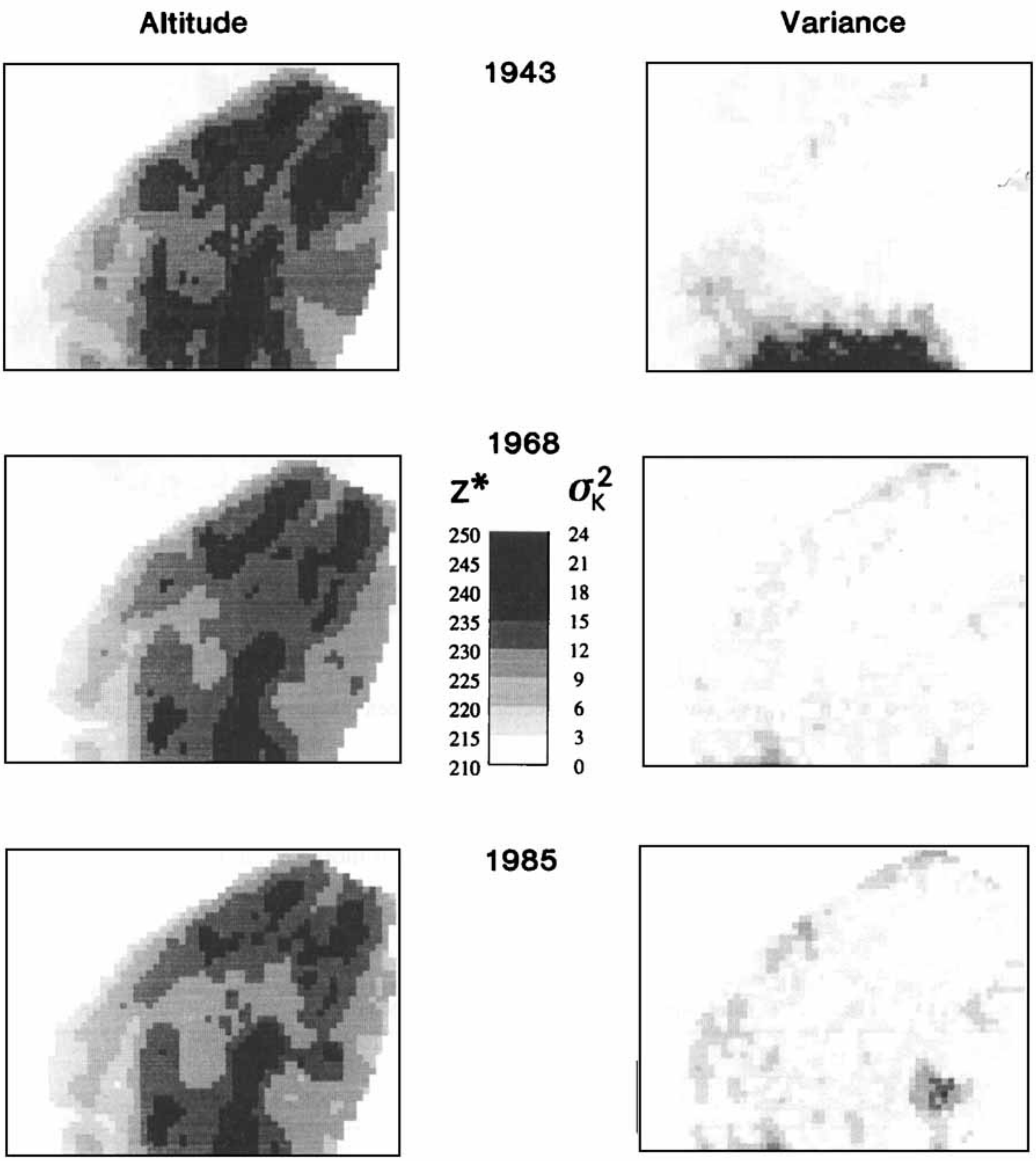

Figure 4. Estimated altitude Z and kriging variance σ_K^2 for the northern part of the moraine complex ('mn') in 1943, 1968 and 1985. Grid size is 10×10 m. Variograms used are 'mn 43×68' and 'mn 68×85' (see Table I)

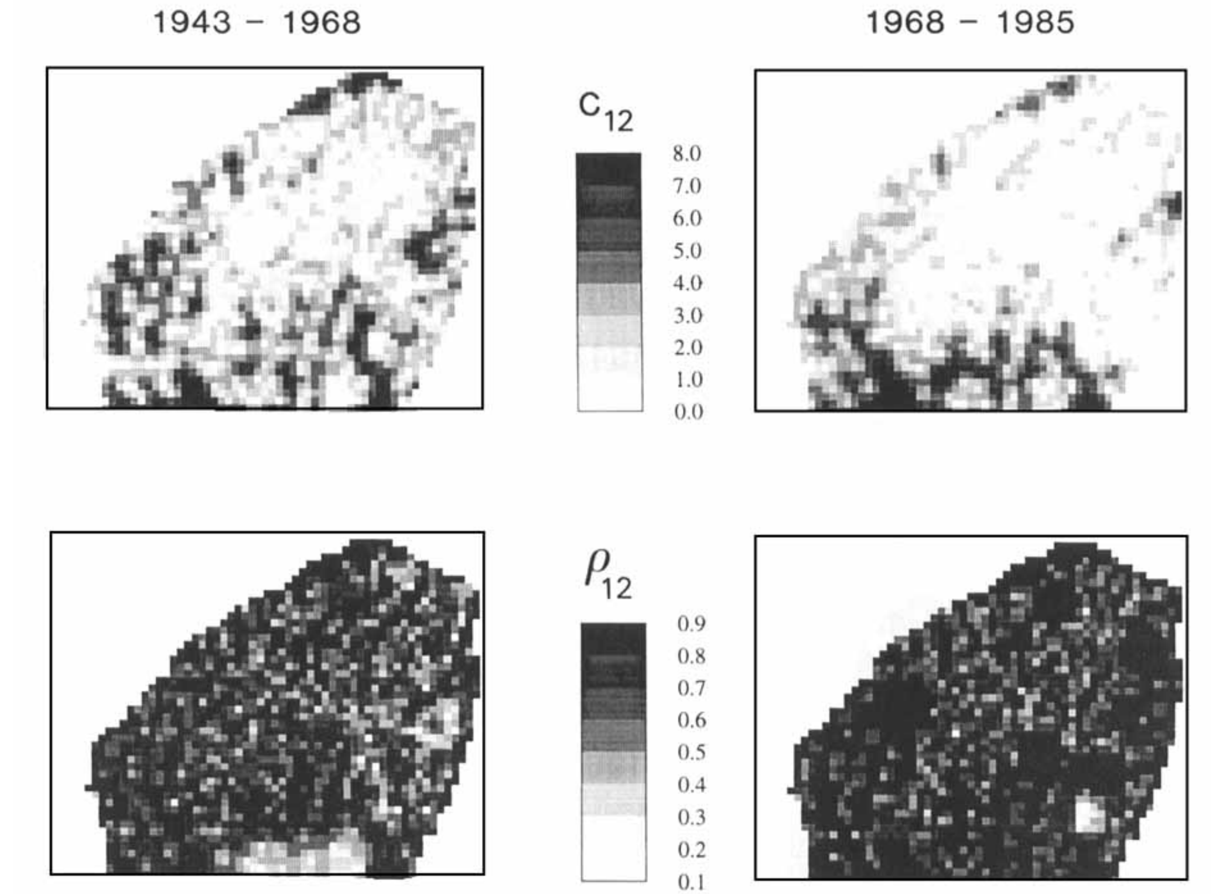

Figure 5. Covariance (C_{12}) and correlation coefficient of interpolation errors (ρ_{12}) for 1943–1968 and 1968–1985, northern part of the moraine complex ('mn'). Grid size is 10 × 10 m. Cross-variograms used are 'mn 43 × 68' and 'mn 68 × 85'

$\sigma^2_{\mathrm{K}}(x)$, and the covariance $C_{ij}(x)$. The correlation coefficient between interpolation errors, $\rho_{ij}(x)$ is given by:

$$\rho_{ij}(x) = \frac{C_{ij}(x)}{\sigma_{\mathrm{K},i}(x)\sigma_{\mathrm{K},j}(x)} \tag{6}$$

Examples of the results are given in Figures 4 and 5. Minimum, maximum and average values for the covariance, correlation coefficient and variance for the northern part of the moraine complex ('mn' in Figure 1) are given in Table II.

Table II. Summary statistics for C_{12} (covariance), ρ_{12} (correlation coefficient) and σ^2_{1-2} (variance) for the northern part of the moraine complex ('mn'). Values apply to the moraine area outlined in Figure 4 (number of grid cells is 1908)

	C_{12}			ρ_{12}			σ^2_{12}		
	min	max	mean	min	max	mean	min	max	mean
1943–1968	−0·07	14·77	2·25	−0·03	0·92	0·67	0·21	23·11	2·69
1968–1985	0·08	13·17	2·81	0·12	0·99	0·78	0·01	15·27	1·43
1943–1985	−0·02	12·77	2·79	−0·02	0·83	0·57	0·32	23·39	4·54

ACCURACY OF ALTITUDE DIFFERENCES

For the analysis of the development of ice-marginal morphology our interest is mainly in altitude differences between subsequent years, given by:

$$dZ_{i-j}(x) = Z_i(x) - Z_j(x) \quad (7)$$

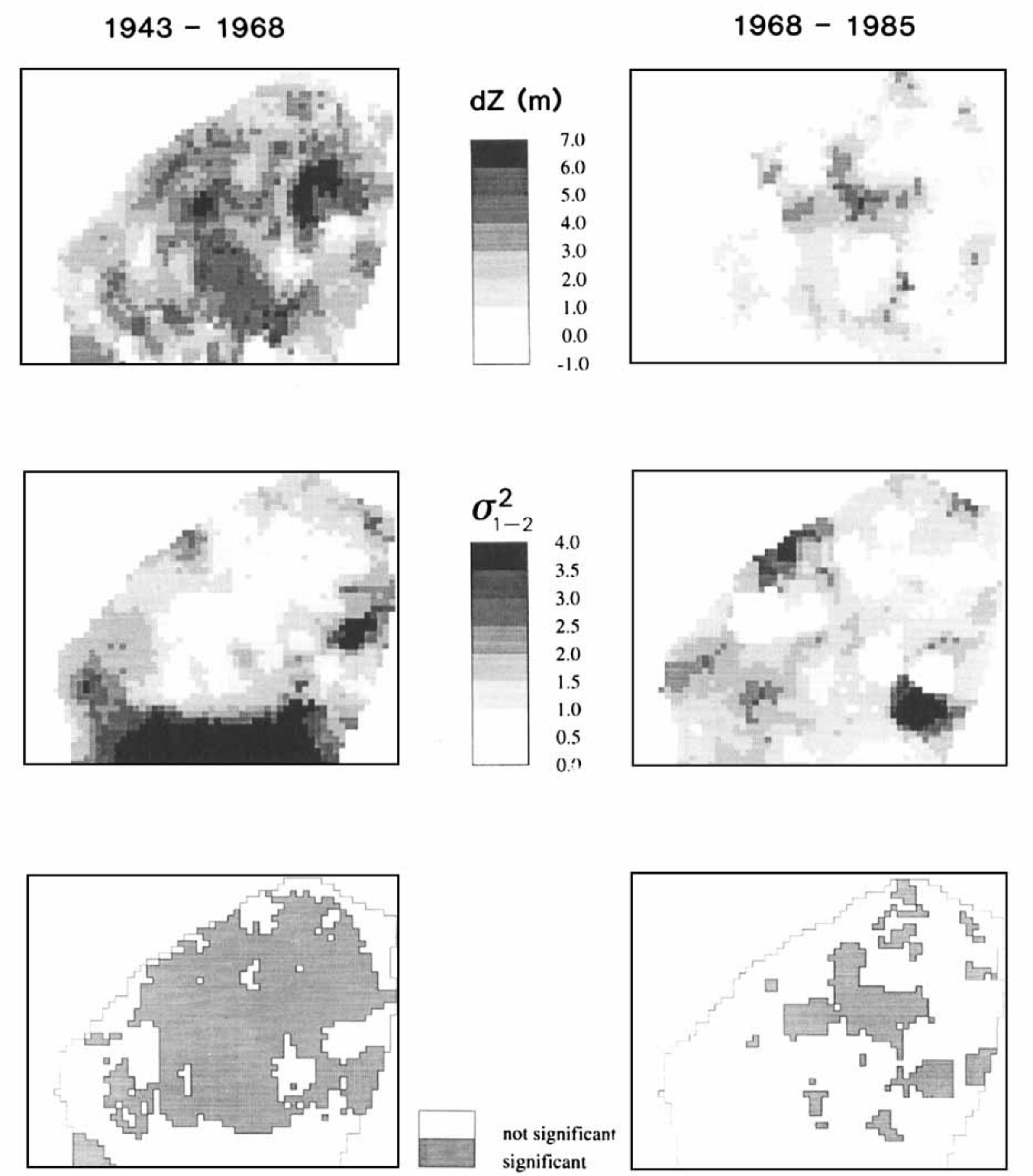

Figure 6. Altitude difference (dZ), variance (σ^2_{1-2}) and areas with significant altitude change for 1943–1968 (period 1) and 1968–1985 (period 2) (northern part of the moraine complex, 'mn'). Gridsize is 10 × 10 m. Positive values of dZ indicate surface lowering

A map of dZ_{i-j} can be estimated by subtracting the kriged maps Z_i and Z_j. The GIS operation (subtracting two maps, years 1 and 2) is linear and therefore the variance of $dZ_{1-2}(x)$ can easily be computed as (Heuvelink, 1993):

$$\sigma^2_{1-2}(x) = \sigma^2_1(x) + \sigma^2_2(x) - 2\rho_{12}(x)\sqrt{\sigma^2_1(x)\sigma^2_2(x)} \tag{8}$$

The last term in Equation 8 illustrates that the relative accuracy depends heavily on the correlation between interpolation errors in subsequent years. Positive correlation causes a reduction of variance. Areas with significant change in altitude dZ are defined as cells where $|dZ| \geqslant 2\sigma_{1-2}$ (Figure 6). The mean altitude change over an area of N grid cells, μ, is calculated from:

$$\mu = \frac{1}{N}\sum_{x=1}^{N} dZ_i(x) \tag{9}$$

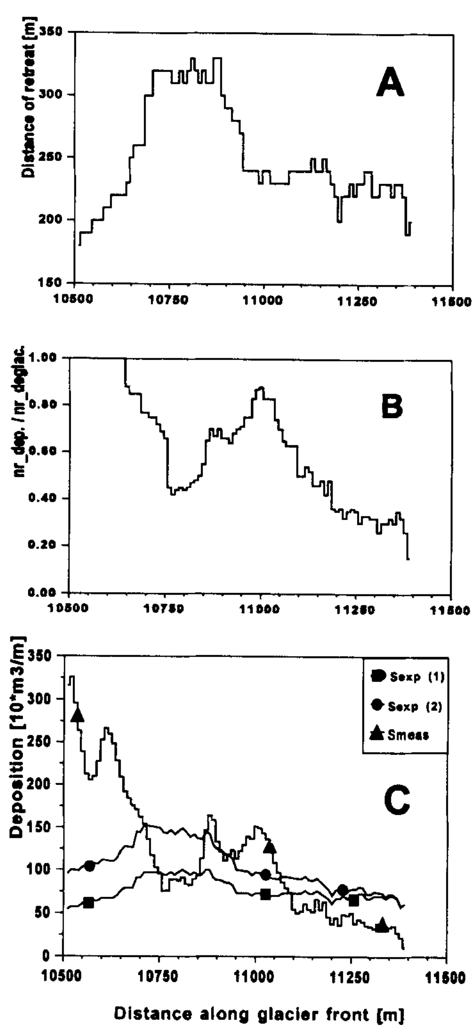

Figure 7. Distance of retreat (A), ratio of cells with significant deposition to number of deglaciated cells (B) and measured, S_{meas}, and expected, S_{exp}, sedimentation (C) for rows 10 m wide perpendicular to the margin of Leverett glacier: 10 500 is south while 11 500 is north (see also Figures 8 and 10D). For $S_{exp}(1)$ a constant basal debris layer of 2·5 m is assumed and for $S_{exp}(2)$ this layer decreases linearly from 5 m in the south to 2·5 m in the north

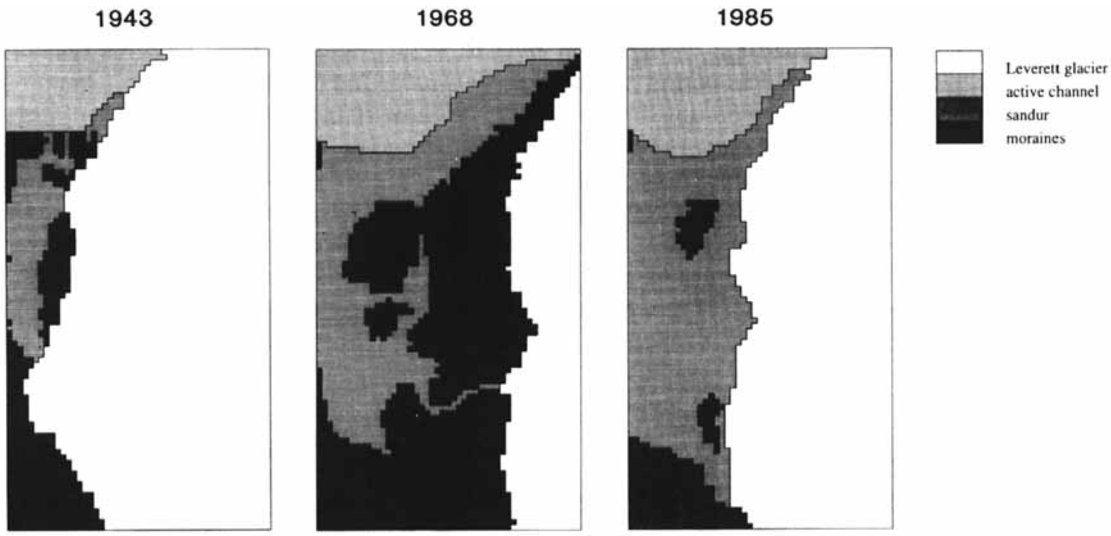

Figure 8. Simplified morphological maps of the proglacial area of Leverett glacier in 1943, 1968 and 1985. Gridsize is 10 × 10 m

where $dZ_i(x)$ is the altitude difference at grid x. The standard deviation of the estimate of the mean, $\delta\mu$ is calculated from:

$$\delta\mu = \frac{1}{N}\sqrt{\sum_{x=1}^{N} \sigma_{1-2}^2(x)} \tag{10}$$

RESULTS: CHANGES RELATED TO DIRECT GLACIER ACTION

A small sandur, enclosed by moraines, was located in front of the 1943 Leverett glacier. The existence of moraines implies that the erosive power of water from outlets in the central and southern part of the glacier

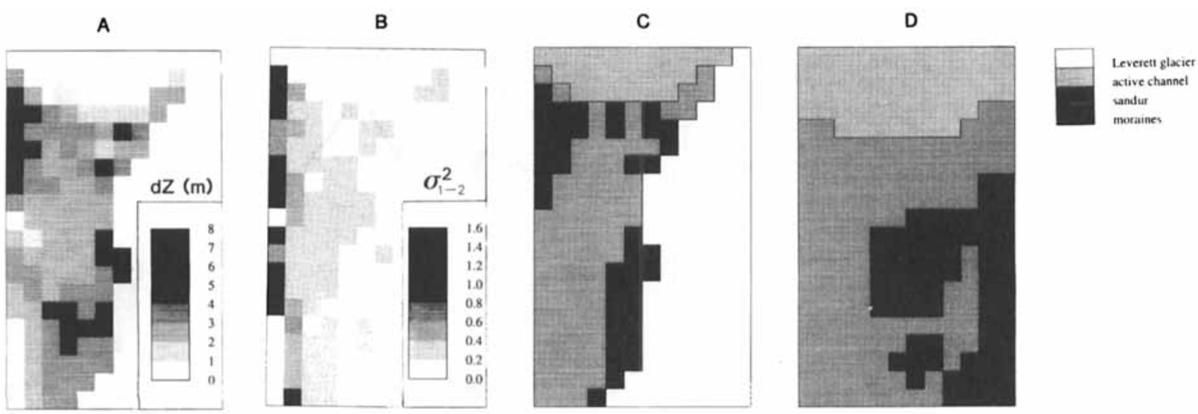

Figure 9. (A) Change in altitude of the sandur in front of the 1943 Leverett glacier. Gridsize is 25 × 25 m. Positive values indicate surface lowering. Only cells with significant change outside the areas occupied by Leverett glacier or active channel are displayed. Variograms used are 'san 43', 'san 68' and 'san 43 × 68' (see Table I). (B) Variance (σ_{1-2}^2); (C) and (D) simplified morphological maps of 1943 and 1968, respectively

front was not high in the period in which Leverett glacier decoupled from the central moraine system. The front of Leverett glacier retreated 180–330 m from 1943 to 1968 (Figure 7; from now on 1943–1968 is referred to as period 1, 1968–1985 as period 2 and 1985–1992 as period 3). During period 1, material was deposited in the area ($0{\cdot}2 \times 10^6$ m^2) occupied by ice in 1943 (Figure 8) and material was eroded in the proglacial area of the 1943 Leverett glacier. The small sandur in front of the 1943 Leverett glacier decreased in altitude by $3{\cdot}2 \pm 0{\cdot}1$ m in period 1, due to the removal of $2{\cdot}1 \pm 0{\cdot}1 \times 10^5$ m^3 of sediment, by the erosion of fluvioglacial sediments and moraines (Figure 9). The morphological development is characterized by the formation of moraines along the glacier front and an increase of the area occupied by fluvioglacial deposits (Figure 8). The latter may be due either to changing meltwater routes or an increase in the amount of meltwater. Figure 10 is the result of 714 measurements in the ice-margin area in the 1968 photographs. In

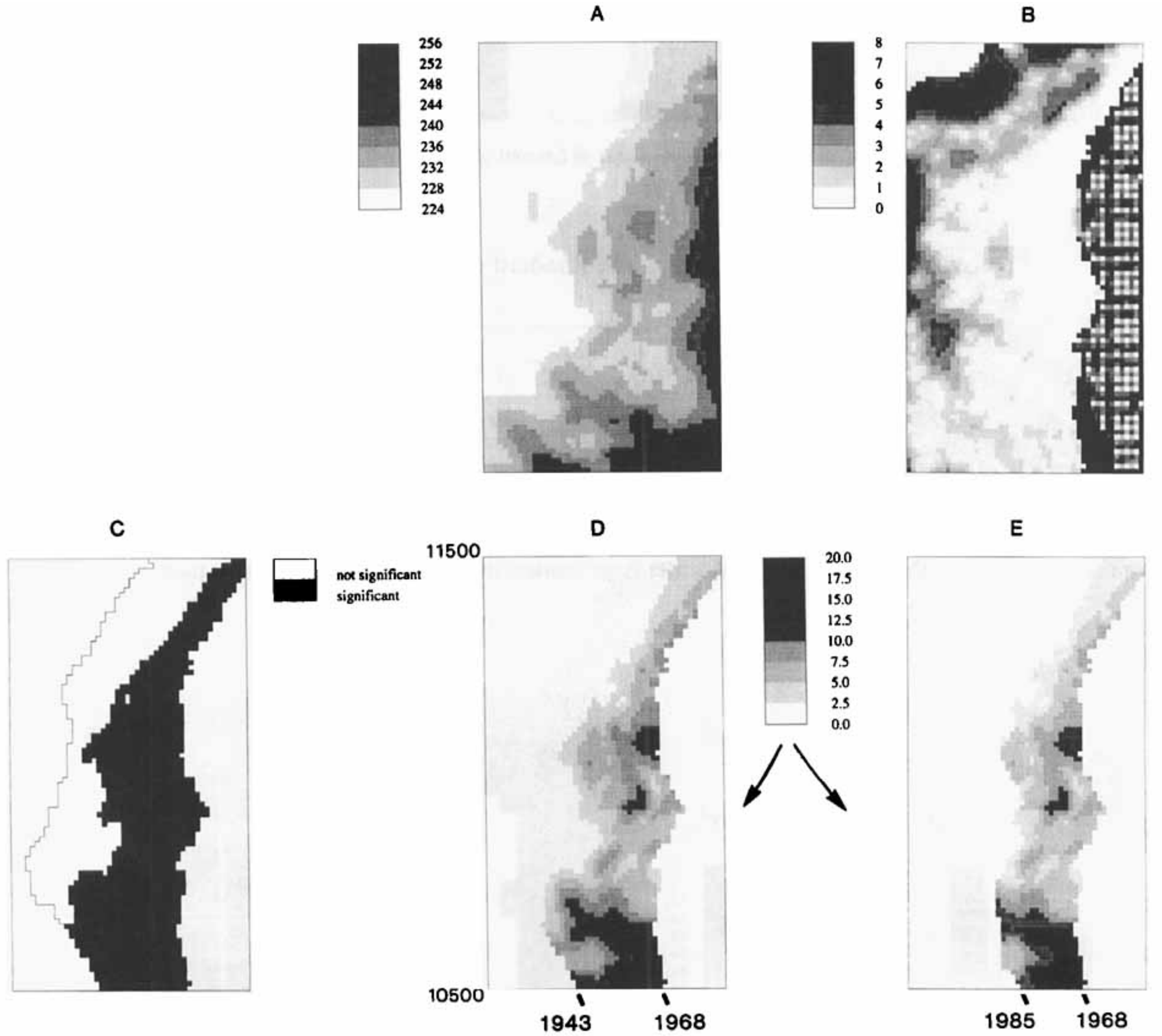

Figure 10. (A) Estimated altitude Z^* and (B) kriging variance σ_K^2 for the ice margin area in 1968 (variogram model 'morsan68', see Table I). Gridsize is 10 × 10 m. Pattern of variance of Leverett glacier is caused by measurements on a regular grid. Ice-margin area and glacier are interpolated separately. (C) Significance of the altitude changes. (D) Material deposited in between the ice margins of 1943 and 1968 (figures indicate the position of the ice margin) assuming a base level of deposition of 226 m. (E) The area overridden by Leverett glacier from 1968 to 1985 (see Figure 8 for morphological maps)

the northern part small moraines are found close to the 1968 ice margin. Moraine volume increases towards the south, interrupted by a frontal channel. In the central part eight ridges can be distinguished from aerial photographs. To estimate the amount of material deposited, a base level of 226 m is assumed (an intermediate altitude of the small sandur) on which deposition took place in period 1.

In Figure 7C volumes deposited are given for rows of 10 m width perpendicular to the ice-sheet margin along the glacier front. Total deposition amounted about $1{\cdot}1 \times 10^6\,m^3$. Not all grid cells have a significant amount of material deposited, because the change in altitude is not always larger than two kriging standard deviations (Figure 10D). The ratio of the number of cells with significant deposition and the number of deglaciated cells is given in Figures 7B and 10C, and shows a similar pattern to the total deposition in Figure 10D. In areas with a small distance of retreat (<250 m, Figure 7A) low deposition is concentrated in a few cells, while high deposition is spread out in all deglaciated cells.

The basal debris layer is the basal zone of the ice sheet and transports most of the debris. If the thickness of the basal debris layer, debris concentration and annual ice-melt volume are known, the 'expected' debris supply (S_{exp}) to a row of grid cells of 10 m width, perpendicular to the ice margin in 1968, is estimated from:

$$S_{exp} = \frac{4}{3} ntD(h_d C_d + h_i C_i)(1 + \phi) \tag{11}$$

With h_d and h_i are the thickness (m) of the basal debris zone and relatively clean ice, and C_d and C_i are their respective debris concentrations, ϕ is the porosity of the accumulated sediment and n the number of deglaciated grid cells. D is the deglaciation rate in metres per year (deglaciation distance divided by deglaciation time t, $t = 25$ years). A factor of 4/3 is introduced to take account of winter advances (it is assumed that the yearly retreat is followed by a winter advance of 1/3 of the retreat distance). Values used are $h_d = 2{\cdot}5$ m, $h_i = 12{\cdot}5$ m, $C_d = 55$ per cent by volume $C_i = 3$ per cent by volume. Values for sediment concentration are based on observations at the present ice margin and within ice-cored moraines. Values for ice thickness are based on the glacier profiles in 1943 and 1968 (Van Tatenhove *et al.*, 1995). Zones with high debris content are concentrated in small bands, an observation also made by Sugden *et al.* (1987) and Knight (1987, 1989) at the glacier margin. Debris concentration within these bands is 55 per cent by volume, which must be regarded as a maximum value, while 3 per cent by volume is the minimum value for relatively clean ice.

The likely spatial and temporal differences in thickness and debris content of the basal debris zone are not taken into account and S_{exp} must be regarded as an absolute figure. Nevertheless, S_{exp} can be used to infer some general patterns of ice-marginal sedimentation. In Figure 7C two scenarios for S_{exp} are illustrated: the first (1) assumes a constant thickness of h_d, and the other (2) assumes a thickness of h_d of 5 m in the south, decreasing linearly to 2·5 m in the north. Both scenarios indicate similar trends:

- The high volumes in the south compared to S_{exp} indicate that the deposition of debris is far from sufficient to produce the observed volumes. The deposited material will therefore be largely (glacier) ice.
- In the northern part S_{exp} exceeds total deposition suggesting syndepositional erosion. Moraines formed close to the main active channel will not have a high preservation potential.
- In the central part S_{exp} is of similar magnitude as measured total deposition. Deposition within eight ridges in the centre provides an average moraine volume of 185 m^3 per metre of ice margin.

Glacier ice remaining behind in period 1 is exposed to melting where the sediment cover is not thick enough to absorb summer heat flux. Around 1968 the glacier started to readvance meeting frontal moraines. Part of the sediment within the frontal moraines is flushed away by meltwater and is temporarily stored in low-angle ice-marginal fans and within the proglacial sandur. The area of this sandur was $1{\cdot}1 \times 10^5\,m^2$ in 1985, and increased in altitude by $2{\cdot}1 \pm 0{\cdot}2$ m from 1968 to 1985. Assuming a base level of glacial erosion of 226 m yields about 20 per cent of the material deposited in period 1, which was temporarily stored in the proglacial sandur after erosion. Some deposited material will be incorporated in the basal zone of the advancing glacier. This would apply especially to ice deposited in period 1. In the proglacial area of the 1985 Leverett glacier only a few remnants of moraines survived the erosion of shifting meltwater channels from central and southern outlets. Prior to 1943, water from these outlets was routed through channels in the southern

part of the moraine complex. At the 1985, 1991 and 1992 glacier margins no moraines were formed despite the large availability of debris. The advancing ice margin incorporated or eroded all previously deposited material.

RESULTS: CHANGES NOT RELATED TO GLACIER CHANGES

Large areas of the investigated moraines are outside the direct influence of the glacier margin. Therefore changes of these moraines cannot be the result of forces directly or indirectly transferred by the glacier. The general pattern in the frontal moraine complex is a significant lowering of surface altitude over large areas during period 1, and a tendency for continued surface lowering at a decreased rate over a smaller area since 1968 (Figure 6).

The northern part of the moraine complex (total area 193 500 m^2, as given in Figure 6) was lowered by $-2{\cdot}90 \pm 0{\cdot}04$ m. The average altitude decrease in the area with significant changes was $-3{\cdot}63 \pm 0{\cdot}03$ m (Figure 6). In period 2, the lowering of the northern part of the moraine complex was $1{\cdot}40 \pm 0{\cdot}03$ m. During period 2, only 315 cells were significantly lowered, of which 63 cells can be explained by fluvial erosion on the east flank of the moraine complex. The average lowering of the remaining 272 grid cells is $-2{\cdot}69 \pm 0{\cdot}1$ m. A large number of the lowered cells in this period were occupied by lakes in 1985 (67 out of 272). The surface area of these lakes increased from 8100 m^2 in 1943, to 8900 m^2 in 1968. In 1985 this area was 9200 m^2. The subsidence of areas occupied by lakes is greater than the average lowering in period 1 ($-4{\cdot}0 \pm 0{\cdot}1$ m) and in period 2 ($-2{\cdot}3 \pm 0{\cdot}1$ m). The estimated values for subsidence of lakes must be regarded as minimum values because account has not been taken of sedimentation of material from the lake shores.

In period 2, 42 cells increased in altitude by an average of $+5{\cdot}1 \pm 0{\cdot}1$ m. Most of these cells are found on the east flank of the moraine complex and in relatively low areas in the centre.

In period 3, most of the eight points measured on the moraines did not change significantly in altitude (Figure 11).

The observed pattern of surface lowering on ridges and lows, with a sandy gravelly composition, and the observations of ice-cored moraines in the southern part of the moraine complex point to melting of sub-surface ice as the explanation of the surface lowering. Although aeolian erosion is active, as indicated by the large number of ventifacts on the surface of the moraine complex, it cannot explain the lowering because of the gravelly composition of sediments within the moraine complex in contrast to aeolian erosion of the fine-grained moraines reported from Spitsbergen (Riezebos *et al.*, 1986).

Pattern of altitude change and morphology of ice-cored moraine

Spatial differences of altitude change can be explained by the irregular pattern of ice-core occurrence,

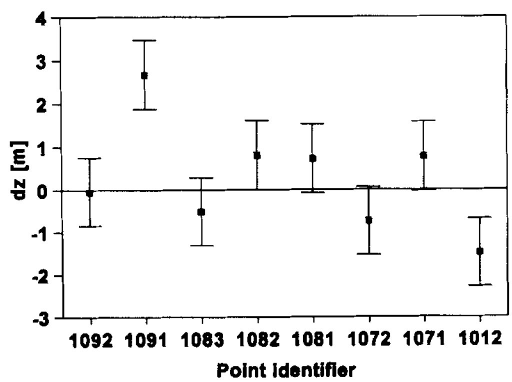

Figure 11. Altitude change of points measured in 1992 on moraines in front of Leverett glacier (for location see Figure 1). Error bars have a magnitude of twice the planimetric error in altitude for the 1985 photographs

variation in soil moisture regimes between ridges and valleys resulting in different active-layer processes, the occurrence of lakes, and drainage patterns and displacement of sediment cover caused by slope instability. Dependent on the morphological position, ridges tend to be drier in summer than valleys. Therefore heat penetration in summer will be more effective in ridges (less pore-ice to melt in the sediment cover) than in valleys. Theoretically this should lead to higher subsidence for cells surrounded by lower cells and low subsidence for cells surrounded by higher cells. In fact local highs and lows do have a higher and lower than average subsidence (Table III). The dependency of altitude change on aspect and slope angle was investigated using the GIS commands *Slope* and *Orient* (Van Deursen and Wesseling, 1992). No significant relationships were found between altitude change and aspect and slope angle (Table III).

The existence of lakes increases the subsidence rate. This is due to more effective heat transfer in summer and winter. In summer, lake-bottom temperature is 7–9°C. Depths of the lakes within the moraine complex are 1–6 m. Regarding the surface area of the larger lakes it is likely that these lakes do not freeze up entirely. Lake-bottom temperatures remain at freezing point during the winter and are therefore relatively high compared to their surroundings. The relatively high subsidence of lakes resulted in relatively steep lake shores, and therefore the decrease in altitude of lake edges is higher than average due to slope failure.

Distribution of water is important in determining active-layer characteristics. Water sources are precipitation, lakes, and water from melting of the ice core. If water refreezes within the moraine system, ice is added which can partly counterbalance surface lowering or even introduce an increase in surface altitude if water supply is large and persistent. The latter process has produced large multi-annual frost mounds on the eastern flank of the moraine complex in period 2. During fieldwork in 1992 only remnants of these mounds were left.

The stability of a sediment cover on thawing slopes is limited and is mainly determined by the thickness

Table III. Altitude change in the northern part of the moraine complex in respect to aspect, slope angle, highs and lows, and lakes. Local highs are cells surrounded by eight cells with a lower altitude, while local lows are surrounded by eight cells with a higher altitude. Aspect, slopes and highs/lows are determined in 1943 for period 1 and in 1968 for period 2; areas occupied by lakes are excluded. Aspect of slopes: north 315–45°; east 45–135°; south 135–225°; and west 225–315°. Altitude change for lakes is calculated using the area occupied by lakes in 1968 and 1985, respectively

	1943–1968		1968–1985	
	n	d*Z*	*n*	d*Z*
Mean	1908	2·90 ± 0·03	1908	1·40 ± 0·03
Aspect				
North	580	2·9 ± 0·1	409	1·1 ± 0·1
East	265	3·0 ± 0·1	184	1·5 ± 0·1
South	517	3·0 ± 0·1	378	1·2 ± 0·1
West	321	2·9 ± 0·1	174	1·5 ± 0·1
Slope angle				
0–4°	307	2·7 ± 0·1	262	1·5 ± 0·2
4–8°	722	3·0 ± 0·1	520	1·2 ± 0·1
8–12°	458	3·1 ± 0·1	274	1·1 ± 0·1
12–16°	156	2·7 ± 0·1	83	1·3 ± 0·1
>16°	40	3·0 ± 0·2	6	1·1 ± 0·4
High/low				
Local highs	9	4·3 ± 0·3	25	1·4 ± 0·2
Local lows	31	2·1 ± 0·3	26	1·5 ± 0·2
Lakes	81	4·0 ± 0·1	92	2·3 ± 0·1

and drainage capacity of the sediment cover (Paul and Eyles, 1990). A general smoothing of relief in time is expected for the ice-cored moraine. Relief looks sharper on 1943 photographs compared with younger photos (an effect partly related to technical differences between the photographs). To evaluate the effect of slope failure on altitude change, it was investigated whether steep slopes (with slope angle larger than 6·4°, being the average slope angle in 1968) have the same position and magnitude as in 1943. In 1943, 1008 gridcells are considered as steep with an average slope angle of 10·0°. Of the 833 steep slopes in 1968,713 (88 per cent) were steep in 1943. The average slope angle is 9·7° ($n = 713$). In 1985, 77 per cent of the steep slopes are at the same position as in 1943 with an average angle of 9·9° ($n = 739$). The minor decrease in slope angle and the fixed position of areas with steep slopes in time point to the relatively unimportant role of slope failure and material displacement by mass movement in determining altitude change, especially compared to ice-core melting. This is probably due to the coarse grain size of the sediment cover of the moraine complex and the arid conditions of the area which prohibit the development of excess pore pressures needed for the decrease of material strength, which can result in slope failure. In the southern part of the moraine complex the situation is clearly different. Here large amounts of silt are incorporated within the ridges and slope instability is much more common, as evidenced by the large amount of silt in slope deposits.

Estimate of the energy used to melt ice cores

The observed changes in altitude in the moraine complex are mainly caused by melting of an ice core underneath a sediment layer. For the analysis of this change average values are calculated for all cells with a significant subsidence and which are not related to other processes causing altitude decrease. These other processes are fluvioglacial erosion on the east and northern flank of the moraine complex and failure of lake shores. Areas occupied by lakes are also excluded. The initial thickness of the melted layer (in m), h_m is given by:

$$h_m = \frac{dZ}{1 - C_d(1 + \phi)} \tag{12}$$

where dZ is the measured altitude change (m), C_d the debris concentration ($<1/(1+\phi)$) and ϕ the porosity of the debris/sediment after melt-out (ϕ is in the range of 0·14–0·46 with an average of 0·3). The altitude change can be regarded as the sum of the melted column containing a certain amount of debris and the settlement of melt-out residue. The thickness of debris melted out (in m) is:

$$h_s = h_m C_d(1 + \phi) \tag{13}$$

The altitude change therefore does not reflect the thickness of melted ice (in m) which is given by:

$$h_i = h_m(1 - C_d) \tag{14}$$

The energy (in $W\,m^{-2}$) required to melt this amount of ice per m^2 is:

$$E = \frac{h_i L_f \rho_i}{\Delta t} \tag{15}$$

where L_f is the latent heat of fusion ($3{\cdot}35 \times 10^5\ J\,kg^{-1}$), ρ_i is the density of ice ($916\ kg\,m^{-3}$) and Δt the time period (25 and 17 years). Results of thicknesses and required energy for two values of debris concentration (3 and 55 per cent by volume) are given in Table IV. High debris concentration results in a higher amount of ice melted at equal altitude change. In view of the fact that the energy required to heat debris above 0°C is very small compared to the energy used to melt ice, E provides an estimate of the deviation of the surface energy balance in case the isolating sediment layer is in balance with 1943 climate conditions. On the other hand, E is the energy required to adjust an initially thin active layer to 1943–1992 climate conditions (see Discussion).

Values for E ($1{\cdot}0$–$2{\cdot}4\ W\,m^{-2}$) averaged over all significantly changed grid cells are about 50 times larger than the geothermal heat flux. The latter can therefore be excluded as an important energy source, which is even more unlikely regarding the results of permafrost studies in the area (Van Tatenhove and Olesen, 1994).

Table IV. Initial thickness (h_m), thickness of settled melt-out residue (h_s), melted ice (h_i) and energy required to melt the estimated amount of ice (E) for two values of debris concentration C_d. Areas occupied by lakes are excluded

	1943–1968		1968–1985	
C_d	3 vol%	55 vol%	3 vol%	55 vol%
h_m (m)	$3{\cdot}80 \pm 0{\cdot}04$	$12{\cdot}7 \pm 0{\cdot}1$	$2{\cdot}8 \pm 0{\cdot}1$	$9{\cdot}4 \pm 0{\cdot}1$
h_i (m)	$3{\cdot}70 \pm 0{\cdot}04$	$5{\cdot}70 \pm 0{\cdot}02$	$2{\cdot}7 \pm 0{\cdot}1$	$4{\cdot}2 \pm 0{\cdot}1$
h_s (m)	$0{\cdot}20 \pm 0{\cdot}01$	$9{\cdot}10 \pm 0{\cdot}03$	$0{\cdot}10 \pm 0{\cdot}02$	$6{\cdot}7 \pm 0{\cdot}1$
E (W m^{-2})	$1{\cdot}40 \pm 0{\cdot}02$	$2{\cdot}20 \pm 0{\cdot}02$	$1{\cdot}0 \pm 0{\cdot}1$	$1{\cdot}6 \pm 0{\cdot}1$

From these studies a permafrost thickness of about 100 m is envisaged and it is highly unlikely that the base of the permafrost coincides with the base of the ice core.

DISCUSSION

The retreat of Leverett glacier in period 1 is associated with collapse of lateral moraines, deposition of frontal moraines and erosion of a small proglacial sandur. Along the front of Leverett glacier formation of ice-cored moraines dominates the southern part, while in the norther part moraines without an ice core are formed. This is the reflection of the different role of synsedimentary fluvioglacial erosion/deposition and possibly the glacier thermal regime as well. In the southern part, the glacier margin is likely to be frozen to the subsoil, while the persistent, high discharge carrying channel in the north creates (local) temperate conditions. These results illustrate that different moraine-forming processes take place contemporaneously at one glacier. General landscape models for the development of sedimentary facies of morphology (e.g. Eyles, 1983) should be regarded as helpful hypotheses, but the likely variability in sedimentary products should be kept in mind.

In the central part of the glacier, moraines were formed with an estimated volume of 185 m^3 per metre of ice margin. Even if they are partially ice-cored, the moraine volume is of the same magnitude as the moraines in the area dated by ^{14}C at 6000–7000 a BP. The formation of the Leverett moraines in a short (decades) retreat period in between advances points to the difficulties in using moraines for mass balance reconstructions.

The formation of thick sequences of debris-rich basal ice is envisaged after 1968 when Leveret glacier advanced over previously deposited ice-cored moraines. It implies that reworking, deposition and reincorporation are normal events during short fluctuations of the ice margin. Ice within the basal debris zone reflects this history and can be influenced by all processes which possibly could affect the previously deposited ice mass. One of these processes, melting by energy from the atmosphere, will influence compositional characteristics if refreezing takes place. This possibility should be evaluated especially for advancing glaciers within a periglacial environment when characteristics such as isotopic composition are studied.

The advance of Leverett glacier is associated with erosion of frontal moraines and (temporary) storage of about 20 per cent of the erosion products in a local sandur, which points to many agents of transport prior to a more definitive deposition. It takes little imagination to answer the question of why form analysis of pebbles gives such poor results in separating till facies in these environments (Östmark, 1988).

The development of an ice-cored moraine complex shows the importance of periglacial processes in determining final characteristics of sediments and landforms. Within a permafrost environment, the classical product of disintegration of ice-cored moraines—melt-out till (Boulton, 1972)—will be influenced by escaping and refreezing of meltwater and the formation of ground ice. The observed multi-annual frost mounds and 'stone boils' produce deformation structures on a larger scale and are not *sensu stricto* related to the collapse of the glacier core. Within these environments a periglacial imprint on the microscale is highly likely for all sediments, while locally periglacial processes may influence structures on the macroscale.

The settlement of an ice-cored moraine is a climate-related process. Such a process is potentially useful to determine local climate change from geomorphological changes. Melting subsurface ice requires energy supplied by a downward energy flux from the surface through an active layer, which tends to increase in thickness in time because of melting of debris-rich ice. Problems in retrieving a climate component from altitude change of ice-cored moraines are therefore the unknown initial sediment cover and the non-linear process of ice-core melting when the thickness of the active layer increases. Lowering of the surface altitude can be associated with a climate amelioration, or with an active layer thickness which is not in equilibrium with climatic conditions. Periods of surface stabilization indicate either local climate deterioration or an equilibrium state of the thickness of the active layer, which protects the underlying ice. An increase of altitude within the moraine complex due to an increase in subsurface ice volume by a vapour flux moving to the freezing front is highly unlikely considering the arid conditions in the area.

Melting of subsurface glacier ice results in a decrease in altitude and the formation of a sediment cover. If the sediment cover, released from debris incorporated within the ice, is larger than the estimated depth of thaw penetration ($\approx 2{\cdot}5$ m for sand and gravel) any melting of ice can be attributed to a deviation of the surface energy balance. The formation of a sediment cover will be quicker with a high debris concentration and a large porosity of the melt-out residue. For an average value of debris concentration (8 per cent by volume) the sediment cover in 1943 should have been thicker than $2{\cdot}0$ m to obtain a significant climate-induced component of altitude change. Such an initial sediment cover thickness is close to the depth of thaw penetration, which seems not very likely considering the calculated changes and present-day sediment cover thicknesses. Therefore the largest part of altitude change is attributed to the adjustment of sediment cover thickness to present climate conditions. The smaller change in altitude in period 2 is explained by the increasing thickness of sediment cover capable of absorbing more energy. The explanation of altitude change points to a recent moraine complex, possibly younger than AD 1900.

CONCLUSIONS

This study shows that with the use of photogrammetry, tightly fixed in an accurate ground-based reference system, a coherent, quantitative analysis of changes in ice-marginal morphology is possible. The use of kriging as interpolation method provides well-defined results within the limits of measurement and interpolation errors. The accuracy obtained is sufficient to draw conclusions on amounts of material deposited and eroded and on surface lowering within an ice-cored moraine complex. In addition to altitude measurements, morphological observations from aerial photographs and in the field provided essential information to interpret patterns of altitude change.

In the period 1943–1968 an area of $0{\cdot}2 \times 10^6$ m^2 was deglaciated. Approximately $1{\cdot}1 \times 10^6$ m^3 of material was deposited in this area. The southern part is characterized by ice-cored moraines while in the north moraines without ice core occur. Differences in depositional product reflect the differences in meltwater activity and probably ice-marginal thermal regime. In the central part of the glacier front the measured deposition equals expected deposition indicating that ice-cored moraines are less probable. During the short retreat phase between the turn of the century and the early 1970s, deposition of eight moraine ridges provides an average moraine volume of 185 m^3 m^{-1}. Therefore small moraines can easily be formed by short ice-margin fluctuations caused by variations in local mass balance. The inability to connect these moraines with glacier fluctuations controlled by climate change has implications for the uses of moraines for palaeoclimatic reconstructions. During deglaciation a small proglacial sandur decreased in altitude by $2{\cdot}7 \pm 0{\cdot}1$ m due to more powerful meltwater activity near the glacier front.

From the early 1970s Leverett glacier advanced over a previously deglaciated area. During this advance, small ice-marginal accumulations were incorporated and subsequently eroded by the advancing glacier. Predominantly in the southern part of the glacier front, thick sequences of debris-rich ice were formed by incorporation of ice-cored moraines. Sediment covers of ice cores and other moraines are eroded and mainly stored within the proglacial sandur.

The northern part of an ice-cored moraine complex decreased in altitude by $-2{\cdot}90 \pm 0{\cdot}04$ m from 1943 to 1968. During the period 1968–1985 this lowering was $-1{\cdot}40 \pm 0{\cdot}03$ m. Analysis of the patterns of altitude

change revealed that areas occupied by lakes subside at a higher rate than average. Although some displacement of sediment cover due to slope failure will have taken place, the minor decrease in slope angle and the fixed position of areas with steep slopes in time point to the relatively unimportant role of slope failure and material displacement by mass movement in determining altitude change. The estimated energy used to melt the subsurface ice is $1{\cdot}4$–$2{\cdot}2\,\mathrm{W\,m^{-2}}$ (period 1) and $1{\cdot}0$–$1{\cdot}6\,\mathrm{W\,m^{-2}}$ (period 2). These values are 30–50 times larger than the geothermal heat flux. Patterns of both altitude change and required energy point to melting forced by energy supplied at the surface.

Melting of subsurface ice took place because the sediment cover was thinner than the average depth of thaw penetration. Considering the fact that altitude change in reaction to a thin sediment cover will decrease in time, the age of the moraine complex could well be younger than AD 1900. This result confirms the modern age of the moraine complex as derived from ^{14}C dating.

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